Prognose der Bodenbewegung bei Starkbeben

Arbeitsbericht Phase IV

(01.01.2005 - 31.12.2007)

An diesem Bericht haben mitgewirkt:

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5.1 General Information about the completed project A7

5.1.1 Title:

Strong Ground Motion Assessment

5.1.2 Research Areas:

Seismology

5.1.3 Principal investigator:

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5.2 Project development report

5.2.1 Report

Abstract
An essential part of seismic hazard analysis is the establishment of attenuation functions that allow the computation of ground motion parameters. Attenuation functions for a specific region are usually, in the lack of a large database, either derived from only a few strong motion records, or even simply taken from a different region, that shows in the best case a similar geological and tectonic setting. This procedure leads to attenuation functions that can either over- or underestimate the ground motion parameters. A major issue of the project A7 was to optimize the computation of parameters which are used for stochastic forward modeling of ground motions by integrating not only a few strong motion records but also synthetic seismic records derived from modeling with the Empirical Green’s Functions (EGF) method and a Finite Difference (FD) method. That way we take advantage of additional information we have about the region: our knowledge of seismic records of small events used as empirical Green’s functions as well as our knowledge about the sub-surface structure that was derived from seismic refraction experiments (HAUSER ET AL. 2001 2007) and seismic tomography (MARTIN 2005 2006) and is needed as an input to FD-modeling.

Finite-Differences (FD) Modelling
Ground motions up to 4.5 Hz of the August 30th 1986 and October 27th 2004 earthquakes were simulated by using a 2D FD method (KARRENBACH 1995). The simulation results were compared with recorded seismograms and in the case of the 1986 earthquake also with observed intensities.

Site effects were implemented by applying frequency dependent amplification ratios to the seismograms simulated by FD modeling.

Modelling with Empirical Green’s Functions (EGF)
The EGF method of IRIKURA (1983, 1986, 1999) was used to get more insight into the source parameters of the two moderate earthquakes of October 27th 2004 and May 14th 2005 as well as the large shocks which occurred on August 30th 1986 and March 4th 1977 (OTH ET AL. 2007a,c). Synthetic time series (acceleration, velocity and displacement) were computed for these four Vrancea earthquakes in a broadband frequency range (about 0.4 – 12 Hz) from smaller events using the EGF-method of Irikura and compared with observed records. As there is only one observation available for the 1977 event, instrumentally determined intensity was compared with the
observed macroseismic intensity (MSK) pattern. Minimizing the cost (misfit) between observations and simulations led to suitable source models for these events (strong motion generation area, SMGA, after Miyake et al. 2003). A genetic algorithm was used to find acceptable solutions.

The results from this study, besides new physical insights into the source mechanism of Vrancea earthquakes, provide the necessary information on the scaling characteristics of these earthquakes to generate a large catalogue of synthetic time series from scenario earthquakes. This database comprises more than 160 simulated events with moment magnitudes ranging between 5.5 and 8.0 at 43 K2 stations (Greçu et al. 2004) and fills the gap between the large amount of weak motion recordings acquired with the K2 network in recent years and the few strong motion records from past large Vrancea earthquakes by using the weak motion records to generate data-consistent simulations.

Figure 1: Distribution of strong motion stations in SE Romania with a zoomed view of the Bucharest area (gray inlay). The stars depict the epicenters of the 1986 (white) and 2004 (black) earthquakes.

**Strong-Motion-Catalogue**

In order to estimate realistic ground motion parameters, we included records of strong Vrancea earthquakes in the determination of those parameters (Fig. 1). On August 30, 1986, a $M_W = 7.1$ event occurred in the Vrancea region and was recorded at 20 stations in Romania. In 1990, two strong earthquakes took place in the region (May 30, 1990 – $M_W=6.9$, May 31, 1990 – $M_W=6.4$). On October 27, 2004 a $M_W = 6.0$ event occurred. This event was recorded at 37 stations. In our studies, we mainly considered records of the 1986 and 2004 strong earthquakes.
Least-Squares Inversion Technique (LSI)

In order to compute ground motion parameters, the three data-sets available are merged together in the frequency-domain. Therefore, Fourier amplitude spectra (FAS) of ground motion need to be computed. The ground motion parameters for which the inversion will be performed are then the parameters that define a FAS of an event of given magnitude, namely the corner frequency $f_c$ and parameters of damping as the quality factor $Q$ or $\alpha, \gamma, \kappa$. The FAS are inverted using a Least-Squares Inversion technique (LSI).

Generalized Inversion Technique (GIT)

As the stochastic simulation technique is based on a realistic description of the FAS of ground motion, the FAS of the data recorded from small and moderate Vrancea earthquakes by the K2-network as well as several recordings of three past strong Vrancea earthquakes (August 30, 1986 – $M_W=7.1$, May 30, 1990 – $M_W=6.9$, May 31, 1990 – $M_W=6.4$) were analyzed using the generalized inversion technique (GIT, e.g. CASTRO ET AL. 1990). The FAS were inverted to obtain their source spectra, the whole path attenuation characteristics and site amplification functions. 55 earthquakes and 43 stations were included in the inversion. The resulting spectral ground motion models, in combination with the scaling analysis performed in the framework of the source parameter study with EGF’s, allow an effective description of the FAS from small to large Vrancea earthquakes over the Romanian territory and constitute the basis for stochastic ground motion simulations.

5.2.1.1 State of Knowledge at the Last Application

The Empirical Green’s Functions Summation Method (EGF) was developed by HARTZELL (1987) and WU (1978) and is regarded as a powerful tool for site-dependent ground motion estimation. Empirical scaling relations or the estimation of a source model with realistic strong motion parameters are needed for the step from small earthquakes to large events. Composite-approaches (IRIKURA 1983) use scaling relations that assume a constant stress-drop. Therefore both, small and large earthquakes have to have magnitudes for which the scaling relations are valid. This is typically given if small earthquakes are two magnitudes smaller than the large events to be modelled.

Finite-Differences Methods (FD) are used extensively in the last years for varying scientific questions. OLSEN (1994, 2000) examined effects of sedimentary basins and showed that the 3D geological structure has a strong influence on strong ground motion. Damping has to be considered important and can for example be implied using correction functions. FRANKELL AND STEPHENSON (2000) modelled small earthquakes using a 3D FD method for the Seattle Fault Zone and reproduced successfully peak amplitudes and signal duration for the long-period range
 (> 2 seconds). A hybrid method was, for the same region, applied by HARTZELL ET AL. (2002). The low-frequent part of the signal was simulated using Finite Differences, while the high-frequent part of the signal was simulated with a stochastic method A non-linear correction has to be applied in order to reduce large amplitudes.

Stochastic methods of ground-motion simulation are based on attenuation functions. Attenuation functions have been developed by several authors (e.g. BOORE ET AL. 1997, YOUNGS ET AL. 1997) for different regions of the world. In fair distance from the source, high-frequent ground motion can be described by the phase spectrum of a band-limited Gaussian noise of finite length and the characteristic $\omega^2$ amplitude spectrum of the source. The code FINSIM (BERESNEV AND ATKINSON 1997, 1998) modifies this idea in a way that not only point sources but also ruptures of finite length can be considered.

5.2.1.2 Applied Methods, Results and Their Importance

Methods

The Finite-Differences Method

A 2D FD method (KARRENBACH 1995) was used in order to simulate wave propagation of the Vrancea strong earthquakes up to 4.5 Hz. The FD scheme is fourth order in space and second order in time. To simulate area-wide ground motion parameters of the August 30, 1986 and October 27, 2004 earthquakes, wave propagation was modeled for 20 different slices, which are rotated around the epicenter – hypocenter axis (Fig. 2).

As it is not possible to infer the correct 3D amplitudes and waveforms from 2D modeling, we developed a correction method (MIKSAT ET AL. 2008) that translates seismograms simulated in 2D into 3D seismograms.

The used subsurface structure of the crust is mainly based on two refraction seismic lines (HAUSER ET AL. 2001 2007). Additional information comes from 3D refraction tomography (LANDES ET AL. 2004) and receiver function studies (DIEHL ET AL. 2005). Based on these information Martin et al. (2005) compiled a crustal model of the region. The mantle structure was derived by nonlinear teleseismic body wave tomography (MARTIN ET AL. 2005 2006). In order to explain the recorded seismograms, it was necessary to introduce scatterers into the known underground model. Unfortunately, there is no information on the scattering properties for Romania available. Therefore, the stochastic properties are based on HOCK ET AL. (2004). We apply correlation lengths of 2 km an RMS velocity perturbation of 5 % for the crust and 4 km and RMS velocity perturbation of 2 % for the mantel (Fig. 3).

The applied Q structure was adopted from SOKOLOV ET AL. (2004). Q is 150 $f^{0.80}$ for depths greater than 100 km, 400 $f^{0.90}$ between 40 and 100 km and 100 $f^{0.80}$ above
40 km. To take into account the site effects, the seismograms from the FD simulations were amplified by using frequency dependent amplification ratios derived by Sokolov and Bonjer (2006).

The model depth varies depending on the hypocenter depth from 98 km to 131 km. Horizontal extension is 350 km and the grid discretization is 140 m.

Figure 2: Wave propagation of the Vrancea earthquakes is modeled with a 2D FD method for different 2D slices through the underground structure. The slices are rotated around the epicenter-hypocenter axis. This 2.5D modeling procedure generates area-wide seismograms depending on the number of 2D slices. The figure shows the main crustal features, the epicenter and hypocenter of the 1986 August 30 earthquake and wave propagation for three 2D slices.

Figure 3: 2D slice through station CFR and the epicenter of the smooth model after Martin et al. (2005, 2006) and through the model with added stochastic velocity perturbations.
Source Parameters and Scaling of Vrancea Earthquakes with Empirical Green's Functions

As described by Ott ET AL. (2007a,c), the source characteristics have been investigated for two moderate and two large Vrancea earthquakes (March 4, 1977, $M_W=7.4$; August 30, 1986, $M_W=7.1$; October 27, 2004, $M_W=5.8$; May 14, 2005, $M_W=5.2$, TARGET events). The results were then utilized to calibrate the scaling relations needed for the generation of a catalogue of EGF simulations for hypothetical Vrancea earthquakes with magnitudes ranging between 5.5 and 8.0. This procedure ensures the best possible compatibility between this catalogue and the observations from past earthquakes.

Acceleration data from six EGF events ($4.0 \leq M_W \leq 5.0$) were used to model the 1986, 2004 and 2005 TARGET earthquakes. The 2004 event was itself used as EGF in order to simulate the 1977 TARGET earthquake. The EGF earthquakes were chosen following the conditions that the focal mechanisms should be as similar as possible to the one of the TARGET and they should be located at approximately the same depth.

Figure 4: Topographic map of the Carpathian area. The Vrancea seismogenic zone is situated in the bend of the arc. The epicenters of the utilized earthquakes are marked by stars (big stars for the TARGET events) and the fault plane solutions of the events are also indicated (the corresponding EGF earthquakes are lined up in a column with the respective TARGET event). The K2 accelerometers which provided data for this study are depicted as inverse triangles.
Fig. 4 shows the epicenters and the locations of the stations. Additionally, the focal mechanisms of all earthquakes are displayed. Each analyzed TARGET earthquake is marked by a large star, and the focal mechanisms of the EGF event(s) associated with it are lined up in a column with the TARGET’s focal mechanism. The fault plane solutions of the main shocks are taken from the Harvard CMT catalogue, whereas those of the EGF earthquakes are from the ROMPLUS catalogue (ONCESCU ET AL. 1999b). Concerning the 1977 earthquake, acceleration data from the October 2004 event were used as EGF data. For this earthquake, due to a lack of waveform data (only one recording of the 1977 earthquake in the city of Bucharest), a novel approach was adopted to derive a suitable source model, which consists in using macroseismic intensity to evaluate whether a given source model leads to reasonable simulations or not (OTH ET AL. 2007a,c).

Irikura’s method is based on the self-similarity hypothesis, which in general assumes constant stress drop over a wide magnitude range. Detailed descriptions of the technique are given in IRIKURA (1983, 1986, 1999), MIYAKE ET AL. (2003) and OTH ET AL. (2007a). The source model on which the methodology is based is an extended area with homogeneous slip and rise time (e.g. KAMAE AND IRIKURA 1998; MIYAKE ET AL. 2003). A physical interpretation for this source model has been given recently by MIYAKE ET AL. (2003) (calling this source model the ‘strong motion generation area’, SMGA), who, following the analysis of twelve crustal earthquakes in Japan, came to the conclusion that the SMGA is equivalent to an asperity within a larger rupture area, where the background slip area shows practically no stress release.

The optimal values for the simulation parameters were derived using a genetic algorithm optimization procedure (OTH ET AL. 2007a). For the 1986, 2004 and 2005 earthquakes, the fit between the observed and synthesized acceleration envelope and displacement waveform was used as a criterion of fit. For the 1977 earthquake, simulations were computed using the 2004 event as EGF, intensity estimated from the FAS of these synthetic records using the technique of SOKOLOV (2002) and the residuals between so-modeled and observed macroseismic intensities minimized.

The procedure described led to source models for these four earthquakes which can explain the observed strong motion time histories and the macroseismic intensity pattern of past Vrancea earthquakes. The largest dataset ever used for such a kind of study in Romania has been analyzed.
**Inversion Using the Least Squares (LSI) Method**

The amplitude $A(f)$ of Fourier spectra computed for the horizontal component of seismic data can be expressed as

$$A(F) = C \cdot S(f) \cdot D(f) \cdot P(f)$$  \hspace{1cm} (1)

where $C$ is a constant, $S(f)$ is the source spectrum and $D(f)$ and $P(f)$ are spectral attenuation functions. We compute the natural logarithm of equation (1) and linearise the equation using Taylor’s expansion receiving equation (6). It depends on 5 independent variables and reads like

$$G(f_c, \kappa, \alpha, Q, \gamma) = G_0 + B \cdot \Delta f_c + C \cdot \Delta \kappa + D \cdot \Delta \alpha + E \cdot \Delta Q + F \cdot \Delta \gamma$$  \hspace{1cm} (2)

Here, $G_0$ is $G(f_{c0}, \kappa_0, \alpha_0, Q_0, \gamma_0)$, $\Delta x = (x-x_0)/x_0$ and $f_{c0}$, $\kappa_0$, $\alpha_0$, $Q_0$, and $\gamma_0$ are the starting values for the inversion given below. Thus, the independent variables which are inverted for are the corner frequency $f_c$, the high-frequency-filter $\kappa$, and the factors $\alpha$, $Q(f)$, and $\gamma$, that control the attenuation function $D(f)$. $B$, $C$, $D$, $E$, and $F$ are dependent on the frequency $f$ and the hypocentral distance $R$. Furthermore they depend on the starting values for the inversion $f_{c0} - \gamma_0$ which are compiled in Table 1.

<table>
<thead>
<tr>
<th>Starting values for least-squares inversion</th>
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<tbody>
<tr>
<td>$f_{c0}$ (Hz)</td>
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<tr>
<td>$\kappa$</td>
</tr>
<tr>
<td>$\alpha$</td>
</tr>
<tr>
<td>$Q$</td>
</tr>
<tr>
<td>$\gamma$</td>
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A more detailed description is given in Appendix 1 (GOTTSCHEMMER ET AL., 2006)

**Inversion for Spectral Ground Motion Models with the Generalized Inversion Technique (GIT)**

The stochastic simulation technique is based on an appropriate description of ground motion Fourier amplitude spectra (FAS) as a combination of source, path and site effects (BOORE 2003). These different contributions were separated using the generalized inversion technique (GIT) (e.g. CASTRO ET AL. 1990; PAROLAI ET AL. 2000, 2004; BINDI ET AL. 2006; Oth 2007). The method is applied in two steps: first, non-parametric, frequency-dependent attenuation functions, describing the decay of the spectral amplitude with distance, were derived. In the second step, the FAS were corrected for the effect of attenuation (with the functions derived in the first step) and the corrected spectra were then separated into their source and site contributions.
The equations describing both steps are given by:

\[ \text{step 1: } \log U(f, r_{ij}) = \log M_i(f) + \log A(f, r_{ij}) \]  

\[ \text{step 2: } \log R_{ij}(f) = \log S_i(f) + \log Z_j(f) , \]  

where \( U(f,r_{ij}) \) denotes the observed FAS (of acceleration, horizontal or vertical component) at frequency \( f \) from source \( i \) recorded at site \( j \) (hypocentral distance \( r_{ij} \)), \( M_i(f) \) is a scaling factor for source \( i \), \( A(f,r_{ij}) \) the non-parametric attenuation function, \( R_{ij}(f)=U(f,r_{ij})/A(f,r_{ij}) \) are the data corrected for attenuation, \( S_i(f) \) is the source term and \( Z_j(f) \) the site amplification. These equations are solved separately at each considered frequency. More details on the method can be found in OTH (2007).

From the source spectra, estimates of the corner (angular) frequency \( \omega_c \) (or stress drop, \( \Delta \sigma \), as these can be linked to each other, BRUNE 1970, 1971) and seismic moment \( M_0 \) can be derived. The frequency-dependent quality factor \( Q_S(f) \) is derived from the non-parametric attenuation functions (e.g. BINDI ET AL. 2006). Expressing the FAS using these parameters allows a simple computation of theoretical spectra and hence also the simulation using the stochastic method.

More than 850 records of 55 earthquakes (with magnitudes ranging between 4 and 7, the largest amount of data being restricted to the range \( M_W=4-5 \)) at 43 K2 stations were used (Fig. 5). The FAS of the records were calculated and smoothed around 30 frequency points equidistant in log using the windowing function of KONNO AND OHMACHI (1998). The root-mean-square (rms) average of the two horizontal component spectra was finally obtained and analyzed (the vertical component was also investigated). The dataset, due to the particular geometry, shows several problematic aspects. First, as the sources are very clustered, there are few crossing ray paths from the sources to different stations. Second, due to the large depth of the events (70–200 km), the lowest hypocentral distance in the dataset ranges around 85 km. The distance range of the data (85–320 km) was subdivided into bins 10 km wide and the reference distance was set to \( R_0=90 \) km.

Inhomogeneities in the attenuation characteristics are not averaging out, as all rays from each source to a given station travel similar paths. The traditional approach of describing the whole-path attenuation using one distance-dependent function (at each frequency) for the entire dataset led to unphysical results (OTH 2007). The spectral amplitudes at high frequencies (above approximately 4 Hz) were consistently lower by almost one order of magnitude in region 2 as compared with region 1, an effect which is explainable by inhomogeneities in attenuation, as synthetic data tests revealed (OTH 2007). Therefore, the inversion for the attenuation functions had to be adapted to this special situation. For each of the two regions shown in Fig. 5, a separate attenuation function was derived. The two functions were simultaneously computed in one inversion scheme.
Figure 5: Earthquakes (stars) and stations (inverse triangles) used in the GIT inversion. On the left, a vertical SW-NE cross-section through the epicentral area shows the depth distribution of the analyzed events. The two regions into which the dataset is split for the determination of the attenuation characteristics are also shown.

Only attenuation effects at distances larger than $R_0$ can be evaluated. Therefore, even after correcting the data for the attenuation functions, the spectra $R_{ij}(f)$ are still affected by a cumulative attenuation effect over $R_0$. In the second inversion, separating source and site effects, this residual attenuation is projected into the source contribution, as the site amplification is constrained to be equal to one at the two rock stations MLR and SIR (this constraint is needed to remove the linear dependence between source and site spectra in the second step inversion). In order to remove this remaining attenuation, it is necessary to know at least the value of one source spectrum unaffected by this effect. From the scaling analysis with the empirical Green’s functions method (evaluation of stress drop ratio $C$ and scaling factor $N$ from the spectral ratios, Oth et al. 2007a,c), the source spectrum of the October 27, 2004 ($M_W=5.8$) earthquake is known. Therefore, the residual attenuation over $R_0$ can be corrected from the inverted source spectra.

The results from this study, in combination with the source scaling characteristics derived with empirical Green’s functions, provide a complete description of the effects of Vrancea earthquakes (source, path and site) and contribute to their better physical understanding.
Results

**Finite-Differences Modeling: Simulation of the August 30, 1986 and October 27, 2004 Earthquakes**

We simulated ground motions for the August 30, 1986 and October 27, 2004 earthquakes by combining FD simulation of wave propagation and the known site amplification characteristics of SE Romania (SOKOLOV AND BONJER 2006).

Ten stations, which are located within our model region, recorded the 1986 event. The random velocity fluctuations, which are introduced into the model, produce realistic wave form shapes, but depend strongly on the actual model of random fluctuations beneath the corresponding station. This means that simulation of wave propagation for a set of different random generated fluctuations may produce significantly different waveforms at specific positions. Consequently, it is more appropriate to look at the Fourier amplitude spectra (FAS), which are linked with macroseismic intensity (SOKOLOV 2002) and therefore with damage. We describe the quality of the modeling by comparing modeled and observed FAS. Fig. 6 shows the arithmetic mean of the FAS of the radial and transverse components. Additionally, the reference spectra for intensity VI to X after SOKOLOV (2002) are plotted. The modeled and observed spectra at BUC, CFR, CRL are in good comparison with the observed spectra, which means that the intensity deviation is clearly smaller than one intensity unit. At BAC the modeled spectrum shows slightly smaller amplitudes than the real spectrum. At BAL the modeled values are larger than the observed spectrum. At BAC and BAL the deviation is about one intensity unit. Good comparison is achieved at FOC for frequencies of about 0.5 and 1.5 Hz and for frequencies higher than 3.1 Hz. To give a quantitative measurement of the quality of the modeling, the misfit between observed and modeled spectra of 10 stations is calculated for five frequency ranges (0.11–0.23 Hz, 0.23–0.48 Hz, 0.48–1.02 Hz, 1.02–2.13 Hz, 2.13–4.48 Hz). The misfit for each frequency range is defined as the square residual SR between the logarithmic average observed and modeled spectral values $\log_{10}(S_{\text{obs}})$ and $\log_{10}(S_{\text{mod}})$ within each frequency range:

$$SR = (\log_{10}(S_{\text{obs}}) - \log_{10}(S_{\text{mod}}))^2.$$  

Fig. 7 displays the residuals of all stations at the center frequencies of the considered frequency ranges and the average values for each frequency range. To get an impression of the quality of the modeling, also the SR that would correspond to a deviation of one intensity unit after SOKOLOV (2002) is shown in Fig 7. Next, the synthetic seismograms are translated into macroseismic intensities. This allows an area-wide comparison between modeling and observation and not only a point wise comparison at the location of seismic stations. The relation between FAS and intensities are given by SOKOLOV (2002).
As each intensity value is assigned to a representative frequency range, the minimum intensity which can be evaluated for frequencies up to 4.5 Hz is about VI.

Figure 6: 2.5D modeling of the 1986 earthquake: Modeled and observed Fourier amplitude spectra (FAS) at BAC, BAL, BUC, CFR, CRL and FOC and reference spectra for macroseismic intensities VI to X after Sokolov (2002).

Figure 7: Misfit between the observed and modeled spectra of the 1986 earthquake. Misfit SR is plotted at the center frequencies. The triangles give the average value for each frequency range. The bold black lines correspond to a deviation of one intensity unit.
Figure 8: 2.5D modeling of the 1986 earthquake: The map shows the modeled macroseismic intensities for 20 profiles. The dashed lines indicate the observed isoseismal lines of the 1986 earthquake. A detailed view of the intensities within the dashed rectangle indicated in the top image is shown in the bottom image.

The resulting calculated and observed intensity distributions of the 1986 earthquake are shown in Fig. 8.

Next, we simulated ground motions for the 2004 earthquake. Twelve stations that are located within our model region recorded the event. Fig. 9 displays the misfit SR within the same frequency ranges as for the 1986 earthquake. The average SR values for each frequency range are lower than 0.11, which is about the same as in the modeling of the 1986 earthquake (see Fig. 7).
Only for stations SEC and FUL the misfit SR shows in more than one frequency range larger values than the maximum misfits for the 1986 earthquake. As the representative frequency range of the observed intensities of the 2004 earthquake are larger than the frequency range of the modeling, the SR that corresponds to one intensity value is not plotted in Fig. 9.

Figure 9: Misfit between the observed and modeled spectra of the 2004 earthquake. Misfit SR is plotted at the center frequencies. The triangles give the average value for each frequency range.

Source Parameters and Scaling of Vrancea Earthquakes with Empirical Green’s Functions

As Oth et al. (2007a,c) show, the four analyzed Vrancea earthquakes can be well reproduced using the SMGA source models depicted schematically in Fig. 10 and listed in Table 2. The dimensions and rise times obtained during the inversion for the October 2004 and March 1977 events are remarkably small (around 2 km$^2$ for the former, 65–90 km$^2$ for the latter). For the August 1986 and May 2005 earthquakes, they are somewhat larger (approximately 160 km$^2$ for the former and 12 km$^2$ for the latter). The rise times are very small for all four analyzed Vrancea earthquakes, which directly leads to the conclusion that all of these events show a high particle velocity on the fault and a large static stress drop. From the spectral ratios between the mainshock and each respective EGF event, the stress drop ratio between them was derived and indicates roughly self-similar scaling (Oth et al. 2007a,c).
Figure 10: Sketches of the source models for the four Vrancea earthquakes analyzed. The relative dimensions are scaled correctly and the rupture imitation location is depicted by a star. Note the very similar size of the subfault of the March 1977 and the source model of the October 2004 events.

Table 2: Lowest misfit SMGA models resulting from 5 consecutive runs of the genetic algorithm for each target event with for the 2004, 1986 and 2005 earthquakes. The position of rupture initiation is given as normalized value in the interval [0 1] for the first two events, as these have been inverted using several EGF’s, whereas it is given absolutely for last two (Oth et al. 2007a,c).

<table>
<thead>
<tr>
<th>TARGET</th>
<th>$v_R/v_S$</th>
<th>$L$ [km]</th>
<th>$W$ [km]</th>
<th>$T_r$ [s]</th>
<th>pos. along strike</th>
<th>pos. along dip</th>
</tr>
</thead>
<tbody>
<tr>
<td>October 2004 (M_W=5.8)</td>
<td>0.9</td>
<td>1.16</td>
<td>1.78</td>
<td>0.11</td>
<td>0.3</td>
<td>0.7</td>
</tr>
<tr>
<td>August 1986 (M_W=7.1)</td>
<td>0.7</td>
<td>12.84</td>
<td>12.60</td>
<td>0.26</td>
<td>0.4</td>
<td>1.0</td>
</tr>
<tr>
<td>March 1977 (M_W=7.4, L:W=1:1)</td>
<td>0.9</td>
<td>8.13</td>
<td>8.13</td>
<td>0.96</td>
<td>1</td>
<td>4</td>
</tr>
<tr>
<td>May 2005 (M_W=5.2)</td>
<td>0.9</td>
<td>3.04</td>
<td>3.67</td>
<td>0.08</td>
<td>1</td>
<td>3</td>
</tr>
</tbody>
</table>

Regarding the 1977 earthquake, the lowest misfit SMGA model can explain the observed record at station Incerc (named INB in this study) very well, even though this record was not entering the inversion and hence constitutes an independent piece of information. Moreover, the SMGA size for the 2004 event is in good agreement with the subfault size of the 1977 best model.
Thus, these results are indeed consistent with each other. Radulian et al. (2007) determined similar dimensions for the 2004 event's asperity from the pulse width of the source time function after deconvolution of an empirical Green's function.

The March 1977 as well as the October 2004 earthquakes seem to show 2–3 times larger (static) stress drops than the August 1986 event. Furthermore, all the events analyzed seem to be similar from the dynamic point of view, as they depict almost identical particle velocities and thus, almost identical dynamic stress drops ranging around 1 kbar (Oth et al. 2007a). These large particle velocities are responsible for the strong high-frequency radiation. The stress drops and particle velocities are about one order of magnitude larger than for crustal earthquakes. Miyake et al. (2003) showed that the SMGA is equivalent to an asperity of about 100 bar stress release for crustal earthquakes. The large stress drops and particle velocities should also be taken into consideration when assessing seismic hazard, as they imply a higher energy release than for typical crustal earthquakes.

This study provides both insights into the source physics of the intermediate-depth Vrancea earthquakes, which are inherently different from crustal events, and the scaling relations between small and large earthquakes needed in order to generate a catalogue of synthetic strong motion records of large Vrancea earthquakes.

**Generation of a Catalogue of Hypothetical Strong Vrancea Earthquakes with the EGF method**

Based on the findings described above (Oth et al. 2007a,c), a catalogue of EGF simulations was generated from the same database of small Vrancea earthquakes used for the determination of spectral ground motion models (Fig. 5). The final simulation database comprises 167 events with magnitudes ranging from 5.5 to 8.0.

In Irikura’s technique, the fault plane of the large earthquake is constructed from $N^2$ subfaults of identical size, and $N$ is determined by the following scaling laws (Irikura, 1999):

$$\frac{L}{l} = \frac{W}{w} = \frac{T_r}{t_r} = N$$

and

$$\frac{D}{d} = CN,$$

where $l$, $w$, $d$ and $t_r$ denote the length, width, displacement and rise time of the EGF event and $L$, $W$, $D$ and $T_r$ are the same parameters for the large event. $C$ is the stress drop ratio (dynamic and static) between the mainshock and EGF earthquake.

In terms of seismic moments $M_0$ (mainshock) and $m_0$ (EGF), this means:

$$N = \sqrt[3]{\frac{M_0}{C m_0}}.$$
As shown by OTH ET AL. (2007a,c), the stress drop ratio \( C \) ranges from 0.7 to 2. \( N \) was then derived from the seismic moments using the formula above. The scaling behavior of the other parameters was calibrated with the results from OTH ET AL. (2007a,c). The computed catalogue of synthetics considerably enlarges the available strong motion database from Vrancea earthquakes and is a useful tool in view of seismic hazard assessment.

**Results from the Least Squares Inversion (LSI)**

From the ground motions we compute Fourier amplitude spectra and invert for five unknown parameters \( f_c, \kappa, \alpha, Q, \) and \( \gamma \). Tables 3 and 4 show results for an inversion computed for DS2 and DS3, respectively. The frequency content was restricted for all inversions to the range between 0.5 Hz and 4.5.

The decrease of the corner frequency \( f_c \) with increasing magnitude is in good agreement with theoretical source spectra (e.g. STEIN AND WYSESESSION 2003). However, our values for \( f_c \) are higher than one would expect from the theoretical curves, especially our results for spectra computed with the 2D-FD-method. One possible explanation could be that the rupture time used in the FD-computations was possibly shorter than in reality.

### Table 3: Inversion results DS2 (0.5–4.5 Hz)

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<th>Parameter</th>
<th>( M_W = 5.9 )</th>
<th>( M_W = 7.1 )</th>
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<tr>
<td>( f_c ) (Hz)</td>
<td>16.6</td>
<td>6.4</td>
</tr>
<tr>
<td>( \kappa )</td>
<td>0.09</td>
<td>0.16</td>
</tr>
<tr>
<td>( \alpha )</td>
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<td>0.01</td>
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<tr>
<td>( Q )</td>
<td>5.1e3</td>
<td>8.4e3</td>
</tr>
<tr>
<td>( \gamma )</td>
<td>5.1</td>
<td>2.4</td>
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### Table 4: Inversion results DS3 (0.5–4.5 Hz)

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<tr>
<th>Parameter</th>
<th>( M_W = 5.5 )</th>
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<th>( M_W = 6.0 )</th>
<th>( M_W = 6.5 )</th>
<th>( M_W = 7.0 )</th>
<th>( M_W = 7.5 )</th>
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<tr>
<td>( f_c ) (Hz)</td>
<td>10.1</td>
<td>7.7</td>
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<td>4.3</td>
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<td>1.0</td>
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<td>( \kappa )</td>
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<td>0.01</td>
<td>0.02</td>
<td>0.04</td>
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<tr>
<td>( \alpha )</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
</tr>
<tr>
<td>( Q )</td>
<td>6.3e3</td>
<td>4.4e3</td>
<td>4.0e3</td>
<td>3.5e3</td>
<td>3.2e3</td>
<td>1.0e3</td>
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<tr>
<td>( \gamma )</td>
<td>3.2</td>
<td>2.6</td>
<td>1.9</td>
<td>1.5</td>
<td>1.4</td>
<td>0.2</td>
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Also, large corner frequencies could arise from the fact that it could be due to the fact that the stress drops used in the forward modelling are rather high (e.g. 100 MPa for the 2D FD modelling) in accordance with literature (e.g. SOKOLOV ET AL., 2004).

The $\kappa$ -value takes values of 0.09 ($M_W = 5.9$) and 0.16 ($M_W = 7.1$) for the FD-simulations and ranges from 0.01 ($M_W = 5.5, M_W = 6.0$) to 0.07 ($M_W = 7.1$) for the EGF-simulations. ATKINSON AND SILVA (1997) showed that the parameter should be considered as magnitude-dependent due to the nonlinear behaviour of the rock. For the Vrancea-region, SOKOLOV ET AL. (2004) suggest to use $\kappa = 0.04$ for small and intermediate earthquakes, and $\kappa = 0.08$ for large earthquakes. Thus, our inversion results are in good agreement with this suggestion and confirm it numerically.

Anelastic attenuation as described by eq. 4 comprises factor $\alpha$. There is no obvious magnitude dependence of $\alpha$.

The frequency-dependent damping factor $Q$ shows average values between $1.0e3$ and $8.4e3$. It has be shown that $Q$ decreases with increasing frequency leading to stronger damping for higher frequencies (GOTTSCÄMMER ET AL. 2006).

$\gamma$ decreases slightly with increasing magnitude, apart from the event $M_W = 7.1$. This indicates that ground motion from smaller events experiences higher damping due to geometrical spreading than ground motion from larger events.

*Spectral Ground Motion Models from Weak and Strong Motion Data*

*Attenuation Characteristics*

In order to investigate the attenuation characteristics beneath Vrancea, the first step of the two-step inversion scheme was applied to the spectral amplitudes at each of the 30 selected frequencies. However, the obtained attenuation functions showed a peculiar feature at frequencies higher than about 4 Hz, namely a strong bump (i.e. amplification with distance instead of attenuation), which takes its maximum value at a hypocentral distance of roughly 180 km (Fig. 11, left panel). The reasons for this bump were explored by synthetic data tests described in OTH (2007). The conclusion was that there must be a severe difference in whole-path attenuation properties for stations located in each of the two different regions depicted in Fig. 5. Region 1 comprises the foreland area, whereas region 2 denotes the epicentral area within the Carpathian mountains.
Figure 11: Left: attenuation functions (log A(f, R) versus hypocentral distance R) at four selected frequencies fitting one function to the entire dataset. Note the strong bump at high frequencies. Right: attenuation functions for region 1 (continuous line) and 2 (dashed line). Gray shaded area: ± one standard deviation.

In order to account for these lateral variations in attenuation, two separate attenuation functions were simultaneously inverted for both regions (Fig. 11, right panel). The origin of the function in region 1 at the reference distance R₀ was set to one (respectively zero in log), while the origin of the function in region 2 was free to move with respect to region 1. At low frequencies, the origin is roughly identical in both regions, whereas at high frequencies, a strong offset is visible (about one order of magnitude).

In order to determine the shear wave quality factor Qₘ(f), the obtained attenuation functions were normalized to one at distance R₀, corrected for body wave geometrical spreading (G(R)=R₀/R) and a straight line was fitted to log A₁,2(f, R) plotted versus R (Fig. 12). Thus, only the slope of the functions was evaluated. The obtained Qₘ(f)-values are shown in Fig. 13 (left panel) for both regions. They are similar: Qₘ(f)=114 f⁻₀.₉₆ for region 1 and Qₘ(f)=81 f⁻₁.₁₂ for region 2. The main difference resides in the offset of the origins, not in the slopes of the functions.

A possible explanation for the fact that, apparently, the slopes of the attenuation functions are similar is provided in Fig. 13 (right panel). All the earthquake hypocenters are approximately located almost directly vertically below the stations in region 2. Thus, in order to increase the distance in A₂(f, R), the source depth has to be increased. The difference in travel path for two data points at different distances in region 2 are hence related to the path traveled from the deeper event’s hypocenter up to the one of the shallower one. As schematically indicated in Fig. 13 (right panel), within the gray shaded area, all the rays travel more or less the same path.
Figure 12: Attenuation functions corrected for geometrical spreading (log $A(f,R)$–log $Q(R)$ versus hypocentral distance $R$) and fitted straight line at four selected frequencies for region 1 (continuous line) and 2 (dashed line). Both functions are normalized to 0 (in logarithm) at the reference distance $R_0$ (i.e. same origin), but offset with 0.5 in logarithm in these plots for viewing purposes.

Figure 13: Left panel: derived $Q(f)$ models for region 1 and 2. Right panel: Sketch depicting the ray paths from sources to stations (orientation roughly SW-NE). For the stations in region 2, the rays from different earthquakes all travel the same path through the gray shaded region, which is defined by the depth of the shallowest event in the dataset. Increasing the distance $R$ in the attenuation function for region 2 means to look mainly at data from deeper earthquakes and therefore, only the hypocentral zone is 'sampled'. The offset between the origins of the two attenuation functions is related to a process occurring somewhere in the gray shaded zone.
In conclusion, $A_2(f, R)$ only 'samples' the lower part of the travel path, whereas the large offset observed between $A_1(f, R_0)$ and $A_2(f, R_0)$ is related to some strongly attenuating region somewhere in the gray shaded area, approximately between the shallowest event's hypocenter and the surface.

The seismic tomography results presented by MARTIN ET AL. (2006) and the outcome of the seismic refraction studies (HAUSER ET AL. 2001, 2007), both studies within the framework of the CRC 461, provide indications on the reasons for these strong lateral variations in seismic attenuation. The seismic refraction data (HAUSER ET AL. 2001) suggest the presence of a low-velocity zone at a depth of 47 to 55 km beneath the Vrancea region, which, as SPERNER AND THE CRC 461 TEAM (2005) note, coincides quite well with the observed seismic gap between 40 and 70 km depth already discussed by FUCHS ET AL. (1979). Here, the slab seems to be mechanically decoupled (or only weakly coupled) from the crust through a weak zone. This zone is commonly interpreted as the place where slab detachment currently takes place. This area of weak coupling could be an explanation for the lower spectral amplitudes in the vicinity of the epicentral area (region 2) and the offset between the attenuation functions for the two regions.

Site Amplification and Source Spectra

The spectra were corrected for the effect of attenuation by the functions $A_{1,2}(f, R)$ and inverted in order to retrieve site amplification and source spectra (second step). One constraint (either source or site) is needed in order to remove an undetermined degree of freedom (ANDREWS 1986). As described in OTH (2007), the average of the site amplification at the two rock stations MLR and SIR was constraint to be equal to one.

Some examples of the obtained site amplification functions are shown in Fig. 14. In general, the site functions show large amplification over a wide frequency band, with especially high values at higher frequencies (above about 2–3 Hz). The amplification increases generally with frequency (e.g. SEC, LUC or CER) and stays on a high level, also at frequencies larger than 10 Hz. The analysis was also performed using the vertical component (Z) of ground motion. Here, it is often observed that especially at these very high frequencies, the amplification rises strongly. The maximum amplification on the Z component is generally shifted to higher frequencies than on the horizontal (H) component.
Figure 14: Examples of the obtained site amplification functions in region 1 (H component). Black line: mean of 200 bootstrap samples. Gray shaded area: mean ± one standard deviation.

Regarding the source spectra, an example for the obtained results is shown in Fig. 15 (left panel). The attenuation-corrected spectra still include a cumulative attenuation effect over the reference distance $R_0 = 90$ km. As the constraint used in the inversion is related to the site functions and does not include this attenuation effect, the latter one is moved into the source spectra.

For one of the events in the database, namely the moderate October 2004 event ($M_W = 5.8$), the source spectrum (unaffected by attenuation) is known from the scaling analysis in the EGF study (OTH ET AL. 2007a,c). Therefore, the inverted source spectra can be corrected for this attenuation effect (OTH 2007). The corrected source spectra follow the $\omega^2$-model (BRUNE 1970, 1971) (Fig. 15, right panel) with high corner frequencies, indicating large stress release (in agreement with the EGF study).
Figure 15: Example for the inverted source functions (left) and source functions corrected for attenuation over $R_0=90$ km (right) for an event of magnitude MW=4.1. The obtained corrected source functions follow the $\omega^{-2}$-model (Brune 1970, 1971).

Fig. 16 shows the obtained corner frequencies plotted versus seismic moment (also determined from the spectra). As expected from the scaling analysis in OTH ET AL. (2007a,c), the scaling of the corner frequency approximately follows a self-similar trend, even though there are only few data points available with large seismic moments.

In summary, all the components of ground motion necessary for the description of the FAS (and hence for stochastic simulation) of Vrancea earthquakes have been discussed. The work presented constitutes a complete spectral model, starting from the source spectrum (where, together with the knowledge from the source study with empirical Green’s functions, the corner frequency of any event with a given seismic moment can be computed) via the attenuation characteristics (where the quality factor $Q_S(f)$ has been determined) and ending with the obtained site amplification functions.

Figure 16: Corner frequencies of corrected source spectra versus seismic moment. The fitted straight line has a slope of $-0.3$, which is close to the expected value $-1/3$ in case of self-similarity.
5.2.1.3 Comparison with Research Outside the CRC

SUZUKI AND IWATA (2005) present SMGA parameters from a very similar study than the one performed here with empirical Green’s functions for ten Japanese interplate earthquakes (with depths ranging between 30 and 50 km, which is somewhat shallower than the Vrancea events, but yet larger than for typical crustal earthquakes). These earthquakes show a very similar scaling behavior of the SMGA, although the scatter (especially in the rise time estimates) is also rather large (OTH ET AL. 2007c). Within the uncertainty ranges, the results obtained for the four Vrancea earthquakes can be regarded to be compatible with the ones of SUZUKI AND IWATA (2005). Thus, there is a line of evidence which leads to the conclusion that the scaling behavior for intermediate-depth earthquakes is rather different of the one for crustal earthquakes. The former ones show a much larger particle velocity and static stress drop, and these facts have to be taken into account when performing strong motion simulations and when assessing seismic hazard from this type of earthquakes.

The attenuation characteristics in the Carpathians have also been investigated by other authors. POPA ET AL. (2005), from a rather qualitative analysis of eight small magnitude Vrancea earthquakes which were recorded during the CALIXTO experiment in 1999 (e.g. MARTIN ET AL. 2006), come to the conclusion that in the Transylvanian basin (behind the mountain arc), the epicentral area and the Eastern Carpathians, the spectral amplitudes are lower by up to a factor of 100 compared to those in the foreland platform (region 1 in this work). They also find that the difference in attenuation is much stronger at higher frequencies than at lower ones and that it is most likely not attributable to source or site effects.

RUSSO ET AL. (2005) worked with data from the same network (the accelerometric K2-network). They used 65 small magnitude (mostly smaller than 4) earthquakes recorded in 1999. With several restrictive assumptions (the first one is that the site effect is equal for the S-wave window on the H component and P-wave window on the Z component, and the second one is that the source spectra for P- and S-waves are identical), they used the spectral ratio between the S-wave on the H components and the P-wave on the Z component to derive differential $\delta t^*$ measurements. If a certain relation is assumed between $Q_P$ and $Q_S$, an estimate of $Q_S$ can be derived. Even though there may be some stations with a systematic error due to strong site effects, the problematic aspects mentioned above might average out if one looks at the entire dataset rather than a single station or single $Q_S$ estimates. In summary, they find high attenuation (low $Q_S$) at stations VRI, SIR, OZU and MLR (which are all situated in region 2 defined here) and low attenuation in the foreland (region 1). The results of these two studies are in good agreement with those obtained here.

ONCESCU ET AL. (1999a) used a similar approach as presented here to separate source and site contributions from a (much smaller) dataset of strong (the four large
Vrancea events in 1977, 1986 and 1990) and weak motion (recorded from 1985–1990) spectra from Vrancea earthquakes. They determined a $Q(f)$-model for S-waves ($Q(f) = 109 f^{0.81}$) by using the coda waves from two Vrancea earthquakes at station Incerc in Bucharest (derived with a very small amount of data, but relatively close to the results obtained for region 1 in this study) and corrected the spectra for attenuation and geometrical spreading before performing the inversion. As a site constraint, they used the transfer function calculated from geotechnical data at station Incerc. The different correction of attenuation and the different site constraint make a direct comparison of the results difficult. However, they observe for instance a very strong deamplification at station MLR at high frequencies (also deamplification at VRI), which is most likely due to the fact that their attenuation model is inappropriate for these sites. They did not find strong evidence for nonlinearity by comparing the site functions derived from weak and strong motion data. It is also worth noting that the transfer function which they computed at station Incerc shows a level of amplification quite similar to the amplification function obtained in this work at station INB.

5.2.1.4 Open Questions

The work plan as specified in the application document has been essentially carried out. 3D Modelling and EGF method show promising results, however modelling suffers from a fundamental difficulty as soon as high frequencies are concerned. The (unknown) scattering properties of the earth’s mantle and crust dominate seismograms as soon as the observational distance amounts to many wavelengths – how many depends on the parameters that quantify scattering. As a result, the synthetics look always much simpler than the observations and waveform comparisons become difficult. The improvement of the structural model does not help. This raises the question of what can be modelled: The entire wavefield, e.g. the time series or the Fourier amplitude and phase spectrum or for instance the Fourier amplitude spectrum alone. The quantification of the scattering properties may be a possibility, but detracts from the main purpose of modelling. The conclusion is that

- either significant knowledge on deterministic structural properties of the material within which the waves propagate AND knowledge on the scattering features are required in order to match high-frequency waveforms, or
- hybrid solutions are utilized that combine low-frequency numerical modelling with high-frequency EGFs, or
- only parts of the wavefield (e.g. the Fourier amplitude spectrum) are modelled.

Answers to these questions are sought in a DFG funded project on wavefield modelling for the Taipei Basin in Taiwan.

The original idea of azimuth-depended joint inversion of all three data sets (observed accelerograms, EGF-synthesized accelerograms, FD-modelled accelerograms) did
not turn out very fruitful as the data sets are essentially disjoint, rather than overlapping and complementary. This, however, did not prevent us reaching the scientific goals of A7, as pointed out in the discussion of reported results. Nevertheless, our approach to a better understanding of observed accelerograms by using several methods – empirical analysis and stochastic modelling, Finite Difference modelling, EGF analysis – is most promising and will be used in future work.

We did not manage to work on point D4 of the work plan, where we wanted to include non-linear effects of soft sediments by empirical means. Reason for this is the lack of conclusive information on how much non-linearity is included in the few strong motion records that are available in Romania. The degree of non-linearity remains a matter of debate and is – at least in a way that allows to exploit it for empirical corrections – no acceptable database for the proposed work. Nevertheless, we believe that this remains an important issue as EGF up-scaling and numerical modelling usually suffer from this constraint. Thus this question remains on the research agenda of earthquake engineering in Karlsruhe, where we can exploit the proven co-operation between applied geology, soil mechanics, structural engineering, and geophysics in the future as well.

5.2.1.5 - A7 Reason for Terminating the Project

The project A7 ends due to the completion of the CRC 461.
5.2.1.5 - A7 Literature


5.2.2 List of Publications Resulting from the Project since the Last Proposal

5.2.2.1 Peer-reviewed publications

a) in scientific journals


b) at major scientific conferences


c) in monographs
none

5.2.2.2 Non Peer-Reviewed Publications

a) in scientific journals
none

b) at major scientific conferences


c) in monographs


5.2.2.3 Oral Presentations


5.3 Approval funds for the current funding period

The project was funded in the Collaborative Research Centre from January 2005 to December 2007.

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### 5.3.1 Staffing of the Project

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<th>Name, acad. title, position</th>
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¹ vorher in B3